Neo-Tethys geodynamics and mantle convection: from extension to compression in Africa and a conceptual model for obduction

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Neo-Tethys geodynamics and mantle convection: from extension to compression in Africa and a conceptual model for obduction[1]

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Abstract: Since the Mesozoic, Africa has been under extension with shorter periods of compression associated with obduction of ophiolites on its northern margin. Less frequent than "normal" subduction, obduction is a first order process that remains enigmatic. The closure of the Neo-Tethys Ocean, by the Upper Cretaceous, is characterized by a major obduction event, from the Mediterranean region to the Himalayas, best represented around the Arabian Plate, from Cyprus to Oman. These ophiolites were all emplaced in a short time window in the Late Cretaceous, from ~100 to 75 Ma, on the northern margin of Africa, in a context of compression over large parts of Africa and Europe, across the convergence zone. The scale of this process requires an explanation at the scale of several thousands of kilometres along strike, thus probably involving a large part of the convecting mantle. We suggest that alternating extension and compression in Africa could be explained by switching convection regimes. The extensional situation would correspond to steady-state whole-mantle convection, Africa being carried northward by a large-scale conveyor belt, while compression and obduction would occur when the African slab penetrates the upper–lower mantle transition zone and the African plate accelerates due to increasing plume activity, until full penetration of the Tethys slab in the lower mantle across the 660 km transition zone during a 25 Myr long period. The long-term geological archives on which such scenarios are founded can provide independent time constraints for testing numerical models of mantle convection and slab–plume interactions.

Résumé : Depuis le Mésozoïque, l'Afrique a été en extension avec de courtes périodes de compression associées à l'obduction d'ophiolites sur sa marge nord. Moins fréquente que la subduction, l'obduction est néanmoins un phénomène de premier ordre qui reste énigmatique. La fermeture de la Neo-Téthys au Crétacé supérieur est caractérisée par un épisode majeur d'obduction, depuis la Méditerranée jusqu'à l'Himalaya, en particulier sur la marge de l'Arabie, de Chypre à l'Oman. Ces ophiolites furent toutes mises en place dans un court laps de temps pendant le Crétacé supérieur, de 100 à 75 Ma, dans un contexte de compression enregistré en de larges portions de l'Afrique et de l'Europe, au travers de la zone de convergence. L'échelle de ce processus requiert une explication à l'échelle de plusieurs milliers de kilomètres et donc impliquant vraisemblablement l'ensemble du manteau convectif. Nous suggérons que l'alternance de périodes extensives et compressives en Afrique résulte de changements du régime convectif. Les périodes extensives correspondraient à la convection impliquant tout le manteau, l'Afrique étant portée par une grande cellule de type tapis-roulant, tandis que la compression et l'obduction se produiraient quand le panneau plongeant africain pénètre la transition entre le manteau supérieur et le manteau inférieur et quand la plaque Afrique accélère en conséquence d'une plus grande activité du panache, jusqu'à pénétration complète, durant une période d'environ 25 Myr. Les archives géologiques sur lesquelles ce type de scenarios est fondé peuvent fournir des contraintes temporelles indépendantes pour tester les modèles numériques de convection mantellique et les interactions panache–panneau plongeant.

Introduction

One of the characteristics of mountains formed during closure of the Neo-Tethys Ocean is the existence of large ophiolitic nappes, remnants of oceanic lithosphere (Gass 1968; Coleman 1977, 1981; Dewey 1976). Some of these ophiolites were deeply subducted, metamorphosed, and exhumed to the surface like most of the Alpine ophiolites (Oberhänsli et al. 2004; Lardeaux et al. 2006; Angiboust et al. 2009), whereas other ophiolites were simply thrust over the continental margin, despite an a priori higher density, escaping metamorphism (like the Semail ophiolite in Oman or the Lycian ophiolite in Turkey (Coleman 1981; Okay 1989; Ricou 1971; Şengör and Yilmaz 1981) or again the Newfoundland ophiolites (Kidd et al. 1978; Dewey and Casey 2013)) through a process called obduction. One of the best-documented examples is the Late Cretaceous obduction of the Tethys ocean floor. Two episodes of obduction of oceanic nappes on the northern continental margin of Apulia and Africa are recognized, one in the Late Jurassic – Early Cretaceous in the Dinarides (Dercourt et al. 1986; Schmid et al. 2008) and one in the Late Cretaceous, from Greece to Oman (Ricou 1971; Coleman 1981; Okay and Tüysüz 1999) for which Şengör and Stock (2014) recently coined the name Ayyubid Orogen.

The Neo-Tethys realm is characterized by a second feature: the microcontinent Apulian (or Adria) that separated from Africa sometime in the Jurassic (Ricou 1994; Barrier and Vrielynck 2008; Frizon de Lamotte et al. 2011) and drifted away to collide with Eurasia, forming the Mediterranean belts from the Late Cretaceous onward. At a later stage, Arabia was detached from Africa from the Late Eocene (Jolivet and Faccenna 2000; Bellahsen et al. 2003; McQuarrie et al. 2003). This process of rifting of continental blocks away from the northern margin of Africa also occurred much earlier in the Permian when the Cimmerian blocks rifted away from Africa (Matte 2001) or even earlier in the Devonian when the Paleotethys Ocean formed at the expense of the Rheic Ocean (Stampfli and Borel 2002). Earlier, identical processes were active several times along the Palaeozoic when Avalonia, Armorica, and Hun terranes rifted away from Gondwana (Matte 2001; Stampfli and Borel 2002). We are thus searching for a conceptual model that could explain the following two main features: (*i*) obduction and (*ii*) fragmentation of the northern part of the southern continent and northward travel of the rifted blocks, at the scale of several thousands of kilometres. We herein present a model based on an extensive compilation of geological observations on the obduction process itself and on the large-scale geological evolution of Africa and surrounding mid-ocean ridges.

Obduction of the Neo-Tethys ophiolites

High-pressure and low-temperature (HP–LT) metamorphic conditions recorded in the tectonic units found below ophiolites do not differ from those retrieved in Alpine-type mountain belts and show that obduction results from subduction of the former continental margin below the oceanic lithosphere (Goffé et al. 1988; Searle et al. 1994, 2004; Yamato et al. 2007). Small units made of metamorphic units of oceanic origin, the so-called metamorphic sole, sandwiched between the ophiolite and the subducted margin, sign the first stages of obduction. These metamorphic soles are characterized instead by high-temperature and low-pressure (HT–LP) metamorphic conditions and are systematically older than the HP–LT metamorphic rocks of the subducted continental margin (Hacker 1991; Hacker et al. 1996; Agard et al. 2007).

In Oman or western Turkey, the radiometric ages of magmatic rocks from the obducted ophiolite or the biostratigraphic ages of supra-ophiolite pelagic sediments, such as radiolarites, show that the ophiolite was very young ($<\sim5$ Myr) at the time of obduction (Nicolas 1989; Rioux et al. 2013; Çelik et al. 2006), implying the formation of an oceanic basin along the southern Neo-Tethys margin a few million years before obduction. Two types of models are debated for the Late Cretaceous obduction (Rioux et al. 2013): either subduction initiation along or near a mid-ocean ridge (Nicolas 1989) or formation of intra-oceanic subduction and later oceanic accretion in the upper plate as a result of mantle upwelling (Pearce et al. 1981; Rioux et al. 2013; MacLeod et al. 2013). Whether this young basin was formed as a back-arc basin above a north-dipping subduction, as a prelude to obduction, or along a new mid-ocean ridge is beyond the scope of this paper, but the two models have drastically different consequences. In the latter case, this new subduction would result from the same compressional episode as obduction, but there is then only a short time from the formation of the first intra-oceanic thrust, the formation of a back-arc basin to its obduction. In the former case, the cause for the formation of this new ridge remains enigmatic, but it could be part of the generalized extension that predates compression. In this particular case, the age of the ophiolite should be older than the beginning of compression (i.e., provided by metamorphic sole ages). By contrast, the ophiolite in Armenia was much older at the time of obduction, and published models suggest the existence of an intra-oceanic subduction and a long-lived intra-oceanic back-arc domain where the ophiolite would have been rejenuvated before it was obducted (Rolland et al. 2009; Hässig et al. 2013, 2016).

These different models, obduction starting at the mid-oceanic ridge or as an intra-oceanic subduction associated with back-arc extension (supra-subduction zone ophiolites), provide good explanations for the evolution of metamorphic rocks found below the ophiolites, the HT–LP sole or the HP–LT blueschists and eclogites in the subducted margin (Searle et al. 2004; Yamato et al. 2007), but not for the subduction initiation preluding to obduction.

What is then the cause of the initial thrusting leading to obduction? This question has been approached by two different models: (*i*) acceleration of convergence, as a result of regional-scale plate reorganization (Agard et al. 2006), that makes subduction more difficult below Eurasia and induces transmission of compression within the subducting plate that buckles and finally breaks (Agard et al. 2007); or (*ii*) enhanced compression by increasing mantle plume activity (Vaughan and Scarrow 2003). Open questions include the cause of plate acceleration inducing compression and the cause for the asymmetry of the resulting subduction, always northward (i.e., south-directed obduction) in the case of the Neo-Tethys.

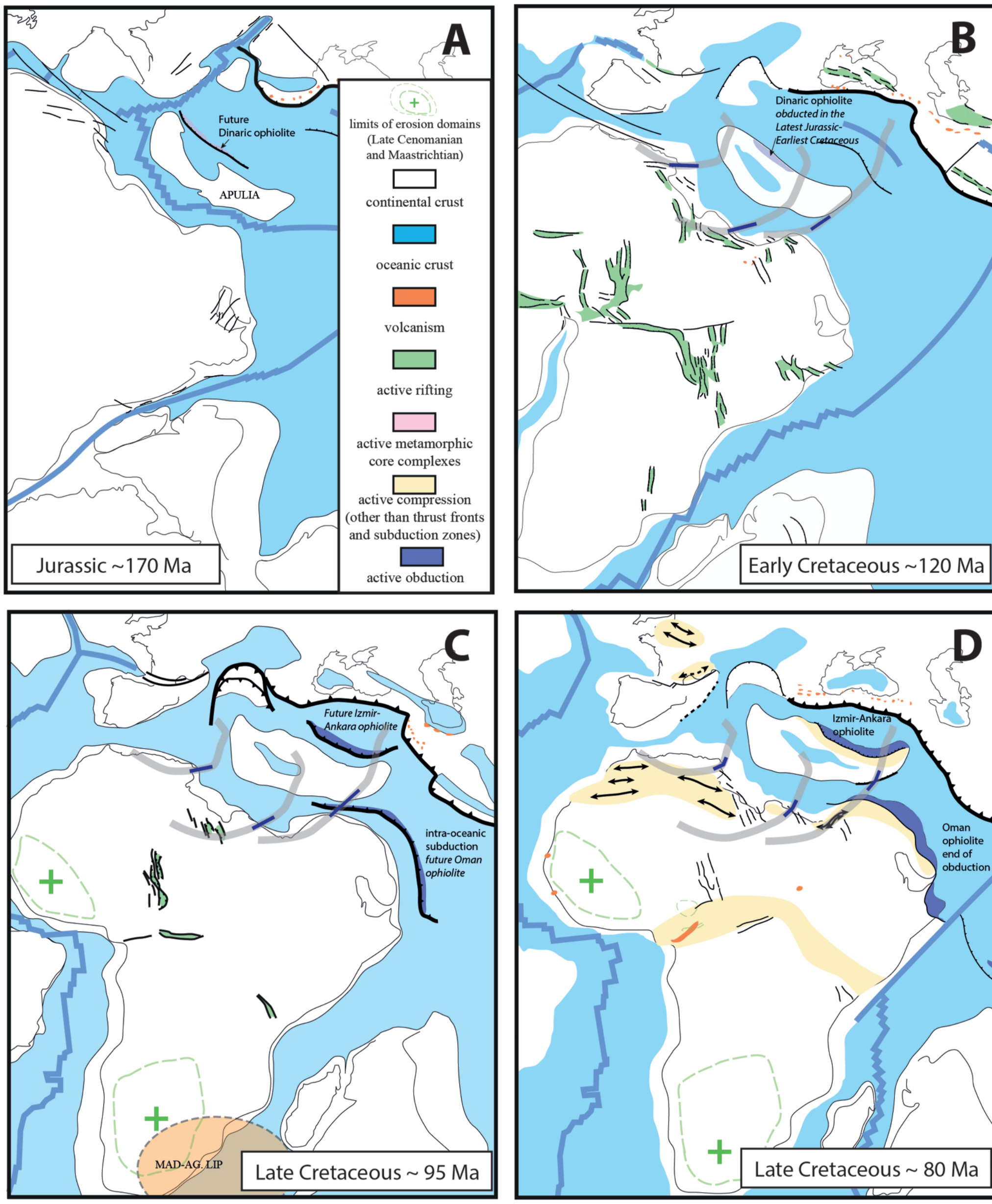
Tectonic history of Africa and the Neo-Tethys Ocean

A series of reconstructions (Fig. 1) and a compilation of geological events (Fig. 2) show the succession of events in the Neo-Tethys Ocean and Africa since the Jurassic. In the Middle Jurassic ($\sim$170 Ma, Fig. 1A), the Neo-Tethys, widely opened in the east, subducts below the southern margin of Eurasia (Agard et al. 2011 and references therein), making its connection with the Atlantic through the Alpine Tethys (Dercourt et al. 1986; Ricou 1994). In the meantime, a second-order rifted basin develops within Africa (Frizon de Lamotte et al. 2011) and the rifted Apulian s.l. continent migrates northward with respect to Africa, forming the eastern Mediterranean Basin (Ricou 1994). The Western Indian Ocean starts to open by rifting between India and Africa during the same period (see a recent review in Frizon de Lamotte et al. 2015). At the end of the Jurassic, the Dinaric ophiolites are emplaced on the northern margin of the Apulian block, as a result of the onset of intra-oceanic subduction dated by metamorphic soles at 175–160 Ma (Agard et al. 2007).

In the Early Cretaceous ($\sim$120 Ma, Fig. 1B), the African Plate is entirely under tension and numerous rifts develop within Africa and Arabia (Guiraud et al. 2005; Frizon de Lamotte et al. 2015), while the South Atlantic Ocean opens. Back-arc basins develop above the northern subduction zone, forming oceanic crust now flooring the Black Sea and the Caspian Sea and small back-arc domains in Central Iran (Hippolyte et al. 2010; Nikishin et al. 1998, 2015*a*, 2015*b*; Agard et al. 2011). Since the Jurassic, an intra-oceanic subduction induces back-arc spreading, forming the ophiolite that will be obducted later on the northern margin of Apulia in Armenia (Hässig et al. 2013, 2016).

After a significant plate reorganization and increase of the Africa–Eurasia convergence velocity at $\sim$118 Ma (Agard et al. 2007), continuing extension on the northern margin of Gondwana and

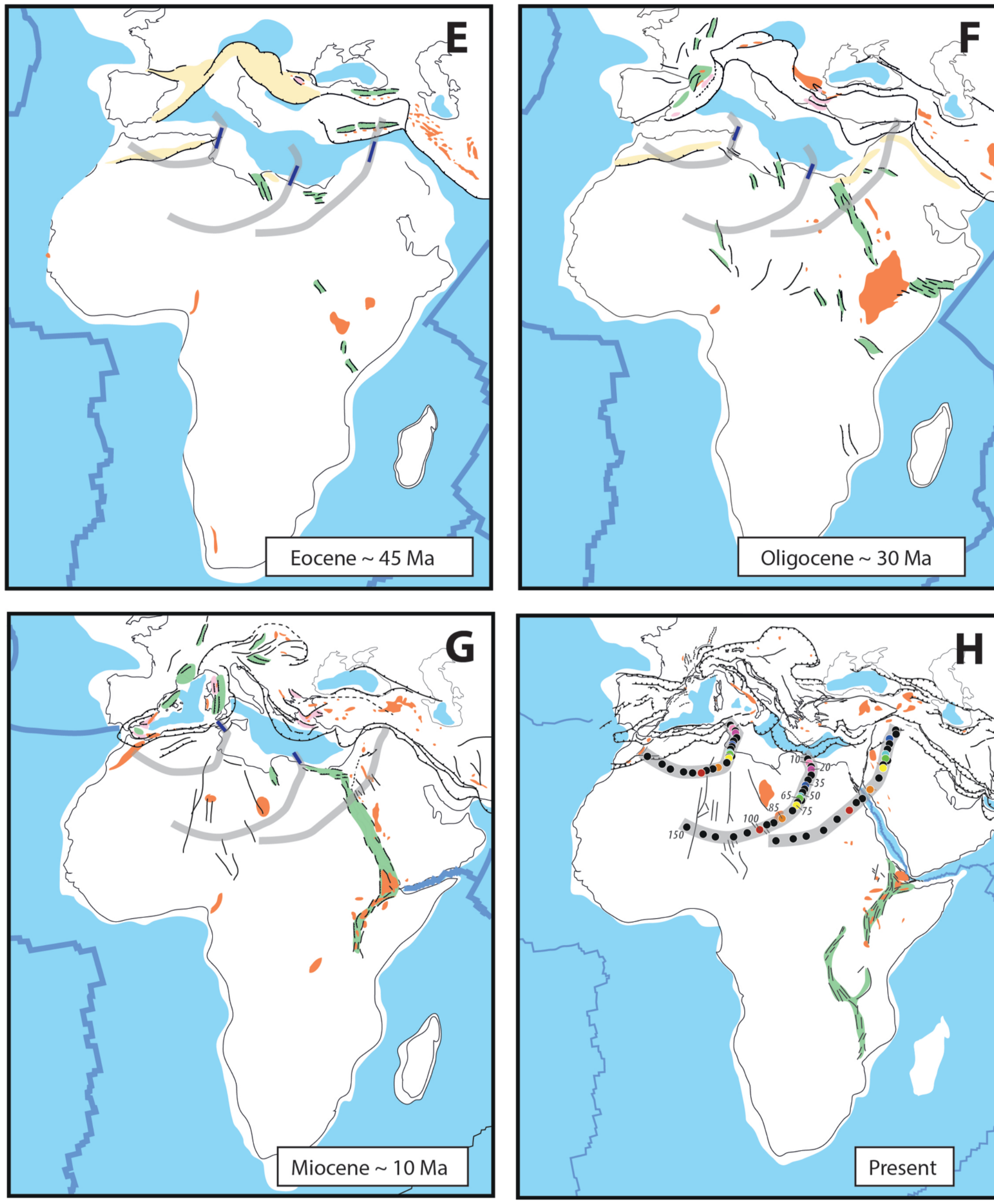
Fig. 1. Reconstructions of Africa and the Tethys Ocean from the Late Jurassic to the Present. Plate kinematics is based upon Barrier and Vrielynck (2008), Jolivet et al. (2003), and Menant et al. (2016). Thick grey lines represent the motion paths of three points of Africa across the reconstructions. The detailed paths with the successive points are shown on Fig. 1H (Present stage) with ages in million years (Ma). Thick blue lines along the paths represent the average direction of motion at the time of reconstruction. Further explanation in text. [Colour online.]



on the active southern margin of Eurasia from 120 to 95 Ma, compression is recorded in Africa and Europe in the Late Cretaceous (Figs. 1C and 1D) (Bosworth et al. 1999; Guiraud et al. 2005). It is preceded by a short event at about 110 Ma characterized by the reactivation of north–south-trending structures (the so-called Austrian phase) mainly observed in western Africa and independent from the Africa–Eurasia convergence. Except for this short earlier event, including the whole obduction process, the compressional episode lasted from ~100 to ~75 Ma. This first compressional event is coeval with the initiation of the intra-oceanic subduction leading to obduction and forming the large ophiolitic nappes observed nowadays (Fig. 1C). The age of this initiation is best constrained by the ages of metamorphic soles that all cluster around 100–95 Ma. Compression then progressively propagates over a large part of Africa with basement undulations and compressional reactivation of the previously formed rifts at about 85 Ma, i.e., the so-called Santonian event (Benkhelil et al. 1988; Benkhelil 1989; Genik 1993; Bosworth et al. 1999; Guiraud et al. 2005; Bevan and Moustafa 2012; Arsenikos et al. 2013) (Fig. 1D). This Santonian event is not present everywhere but is well characterized in the northern part of Africa and Arabia and also along the east–west segment of the Sub-Sahara Rift System. It has apparently not been recorded everywhere and some regions such as the Sirt Basin or the Muglad Rift in Sudan have not been reactivated (McHargue et al. 1992; Wennekers et al. 1996), but it is nevertheless widely distributed all over the northern half of Africa.

Special attention should be payed to the Sirt Basin. This basin developed from the Early Cretaceous to the Present on top of a highly pre-structured and faulted Panafrican basement with a Proterozoic metamorphic basement and a Paleozoic to Early Cenozoic cover (Wennekers et al. 1996; Abadi et al. 2008). Thick accumulations of sediments developed in the Cretaceous with the

Fig. 1 (*concluded*).



irregular deposition of the lower Cretaceous, followed by a period of major syn-rift subsidence during the Cenomanian and Turonian and then by the transgressive deposition of the Late Cretaceous up to the end of the Cretaceous. So, in Sirt, the Late Cretaceous contractional event is only underlined by a decrease of the subsidence rate (Frizon de Lamotte et al. 2011). An additional observation may explain why the Late Cretaceous inversion is not obvious in the Sirt Basin. The main depot-centres of the Sirt Basin strike northwest–southeast, almost perpendicular to the compressional fold axes in the nearby Cyrenaica (Arsenikos et al. 2013). The development of these basins in the Jurassic and Cretaceous was highly influenced by the pre-existing faults and basins with that same strike (Wennekers et al. 1996). The compression recorded in Cyrenaica was not properly oriented to reactivate the Sirt Basin normal faults.

Compression is also recorded during the same period across western Europe, from the Pyrenees where compression starts in the Santonian (~85 Ma, Vergés et al. 2002; Jammes et al. 2010) and southeastern France, all the way to the Paris Basin and the North Sea (Guillocheau et al. 2000), ending up with slow subduction initiation in the western Mediterranean (the future Apennines subduction zone). During the same period, between 95 and 85 Ma, blueschists forming in the subduction zone are exhumed to shallow depth along the southern margin of Eurasia along thousands of kilometres (Agard et al. 2006, 2007; Monié and Agard 2009).

After a period of relative quiescence between 65 and 45 Ma (Rosenbaum et al. 2002), the Middle – Late Eocene shows renewed compression (Fig. 1E) in the Atlas Mountains, Cyrenaica, Syrian Arch, and all the way to the Zagros (Arsenikos et al. 2013; Frizon de Lamotte et al. 2011) and north of the young west Mediterranean subduction zone in the Iberian Range and the Pyrenees (Vergés et al. 2002). This new compressional period preceded the Oligocene uplift of large parts of Africa (Burke et al. 2003; Burke and Gunnell 2008) amounting to 200–300 m in North Africa, coeval with the early formation of the North African volcanic province (Liégeois et al. 2005; Wilson 1993; Wilson and Guiraud 1992). Volcanism is recorded earlier in the Central Sahara, as early as 34 Ma (Ait-Hamou 2006). The uplift is, for instance recorded during the

Fig. 2. Compilation of indicators of mantle convection activity and tectonic and metamorphic events related with the Late Cretaceous compression and obduction event. Green shadings represent the periods of faster Africa–Eurasia convergence, faster spreading in the South Atlantic, and compressional period in Africa in the Late Cretaceous. Labels 1–6: Spreading rates after Conrad and Lithgow-Bertelloni (2007) and Colli et al. (2014). 1: West Indian Ridge. 2: South Atlantic Ridge. 2b: South Atlantic full spreading rate. 3: West Pacific. 4: East Pacific. 5: Global average. 6: Global average with Farallon. Most of the average value originates in the Pacific ridges, especially the peak around 120–110 Ma. Although velocities are much lower, the South Atlantic ridge shows an increased velocity between 120 and 70 Ma. Labels 7–8: Global figure of formation of oceanic crust and sea level variations. The Pacific peak of accretion in the Pacific reflects on the global production rates and the first order high in the sea level is coeval with this globally high production rates in the Cretaceous. 7: Sea-level after Müller et al. (2008*a*, 2008*b*). 8: Global oceanic flux (ridges and plateaus) after Cogné and Humler (2006). Label 9: LIPS after Gaina et al. (2013). Abbreviations: Mad. Agh, Madagascar-Agulhas; SWI, Southwest Indian; SEA, Southeast Atlantic. Labels 10–13: African plate velocity. All models of the Africa–Eurasia convergence show an increase between 120 and 70 Ma that is coeval with the higher velocity along the South Atlantic Ridge. Absolute velocity models show different figures and the peak of velocity is either between 110 and 100 Ma or around 80–70 Ma. 10: Africa–Eurasia convergence velocity at the longitude of the west Mediterranean after Rosenbaum et al. (2002). 11: Africa absolute velocity (Doubrovine et al. 2012; Gaina et al. 2013) based on moving hot spots and a true polar wander model before 124 Ma. 12: Africa RMS velocity, hybrid moving hotspots, and true polar wander corrected reference frame (Zahirovic et al. 2015). 13: Africa–Eurasia convergence velocity at the longitude of Oman (Agard et al. 2007). 14–15: India plate velocity. The velocity peak is recorded after the peak for Africa, around 60 Ma. 14: India absolute velocity (van Hinsbergen et al. 2011). 15: India–Asia convergence (van Hinsbergen et al. 2011). Label 16: Sediments accumulation rates on South African margins (Guillocheau et al. 2012). The peak centered on 80 Ma comes after a period of increasing sediment discharge and a period of uplift and erosion of South Africa. Label 17: Subsidence of northern Africa (Guiraud et al. 2005). Label 18: Timing of subduction, exhumation, and back-arc extension in the western Mediterranean (Jolivet et al. 2003). Label 19: Timing of subduction, exhumation, and back-arc extension in the western Mediterranean (Jolivet and Brun 2010; Jolivet et al. 2003). Label 20: Folding in the Paris Basin (Guillocheau et al. 2000). Label 21: Timing of subduction and obduction in Oman (Agard et al. 2007; Hacker et al. 1996; Nicolas 1989; Rioux et al. 2013). Label 22: Timing of subduction and obduction in Turkey (Çeliket al. 2011; Pourteau et al. 2013). Label 23: Timing of subduction and obduction in Armenia (Hässig et al. 2013, 2016; Rolland et al. 2009). Label 24: Timing of subduction and obduction in the Zagros Internal Zones (Agard et al. 2006, 2005, 2011). Label 25: Timing of subduction and obduction in the Sabzevar Zone in Iran (Rossetti et al. 2014, 2010). Label 26: Timing of subduction and obduction in Sistan (Iran) (Angiboust et al. 2013; Monié and Agard 2009). Label 27: Tectonic timing in the Black Sea region (Hippolyte et al. 2010; Nikishin et al. 2015*a*, 2015*b*). Label 28: Tectonometamorphic timing in the Himalayas (Guillot and Replumaz 2013). [Colour online.]

Early Oligocene in the Niger Delta (Petters 1983), associated with a major regression but its exact age is poorly constrained. Otherwise, the upper plate of the convergence zone to the east records extensional deformation during this period.

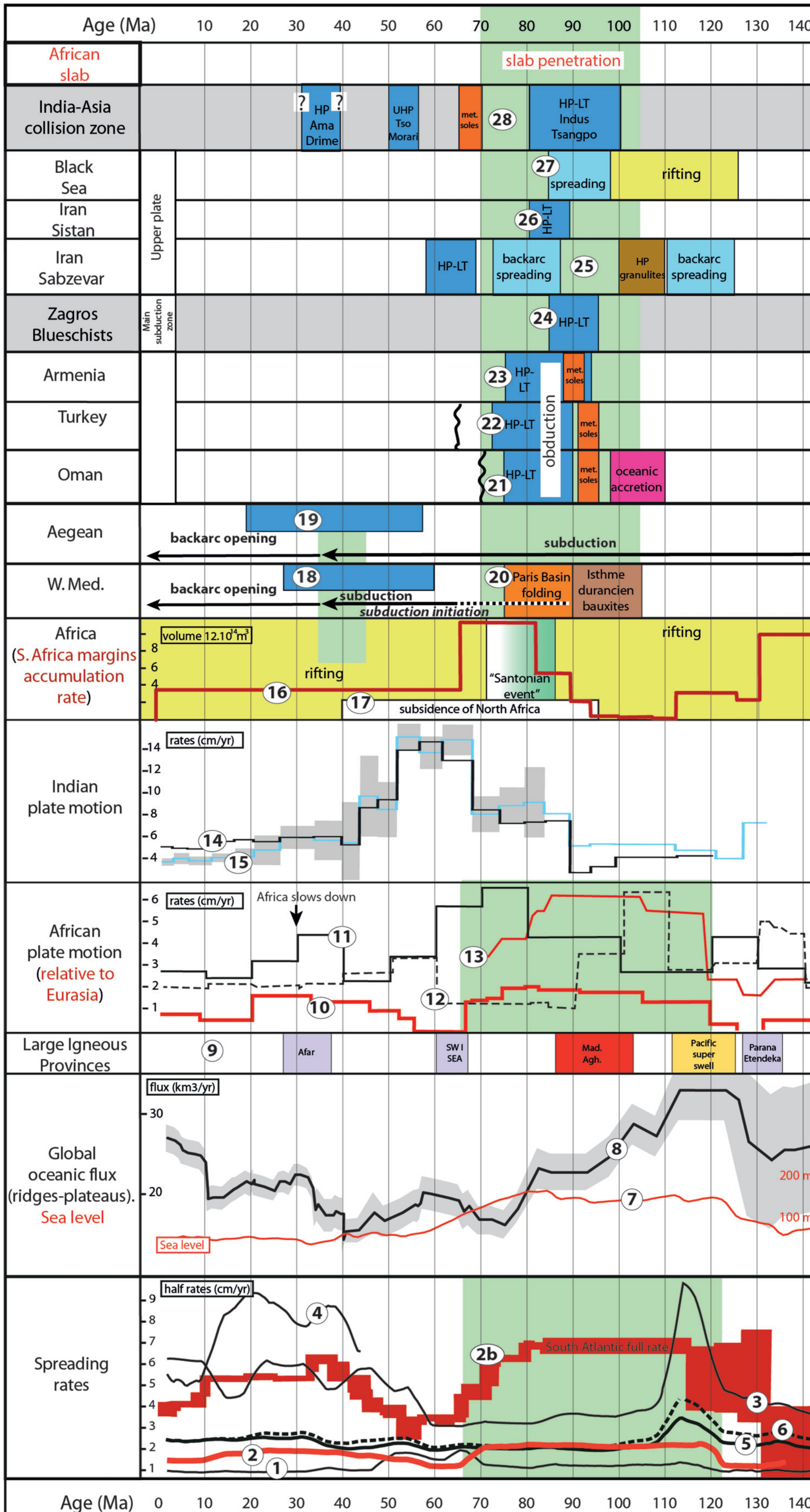
A major change in subduction dynamics then occurs around 30–35 Ma (Fig. 1F), and back-arc basins start to form in the Mediterranean (Jolivet and Faccenna 2000). Meanwhile, rifting starts along the future Gulf of Aden and Red Sea, coeval with the Afar plume-related traps, and most of the sub-Saharan rifts are reactivated. The Miocene and Present stages (Figs. 1G and 1H) are the continuation of the same situation with separation of Arabia from Africa, and opening of Mediterranean back-arc basins.

These reconstructions show that, during 140 Ma, Africa has been mostly under extension and the subduction zone north of it has been forming back-arc basins, except during two periods: (*i*) an ~25 Myr long period (100–75 Ma) of compression associated with obduction, propagating away from the obduction zone within Africa and Europe in the Late Cretaceous and culminating in the Santonian and, after a period of quiescence, from 65 to 45 Ma; and (*ii*) compression resuming at 45 Ma and persisting until ~35 Ma, mostly in the west, before extension was again the predominant regime in Africa and Arabia, except within the Arabia–Eurasia collision zone (Agard et al. 2011; Mouthereau et al. 2012). The intervening extensional periods were associated with plate fragmentation and the formation of Apulia and Arabia. The compressional periods thus seem accidental interruptions in a continuous process of extension and fragmentation of Africa during its motion toward Eurasia. During the Mesozoic this succession of extension and compression and obduction thus occurred at least twice: (*i*) rifting of Adriaand Apulia away from Africa around 180–170 Ma, followed by the Dinaric obduction between 170 and 150 Ma; and (*ii*) distributed extension in Africa in the Early Cretaceous followed by compression and obduction in the Late Cretaceous. The separation of Arabia from Africa from the Late Eocene onwards seems the repetition of the same process, but very little oceanic lithosphere is left and collision occurs coevally (Jolivet and Faccenna 2000).

One important additional observation is that the newly created subduction zones leading to obduction dipped everywhere toward the north, with the same orientation as the already existing subduction zones beneath Eurasia, resulting in a totally asymmetrical situation where the continental lithosphere subducted northward below oceanic lithosphere, while the classical view would be that the oceanic lithosphere sinks below the continental lithosphere, whatever the polarity of subduction.

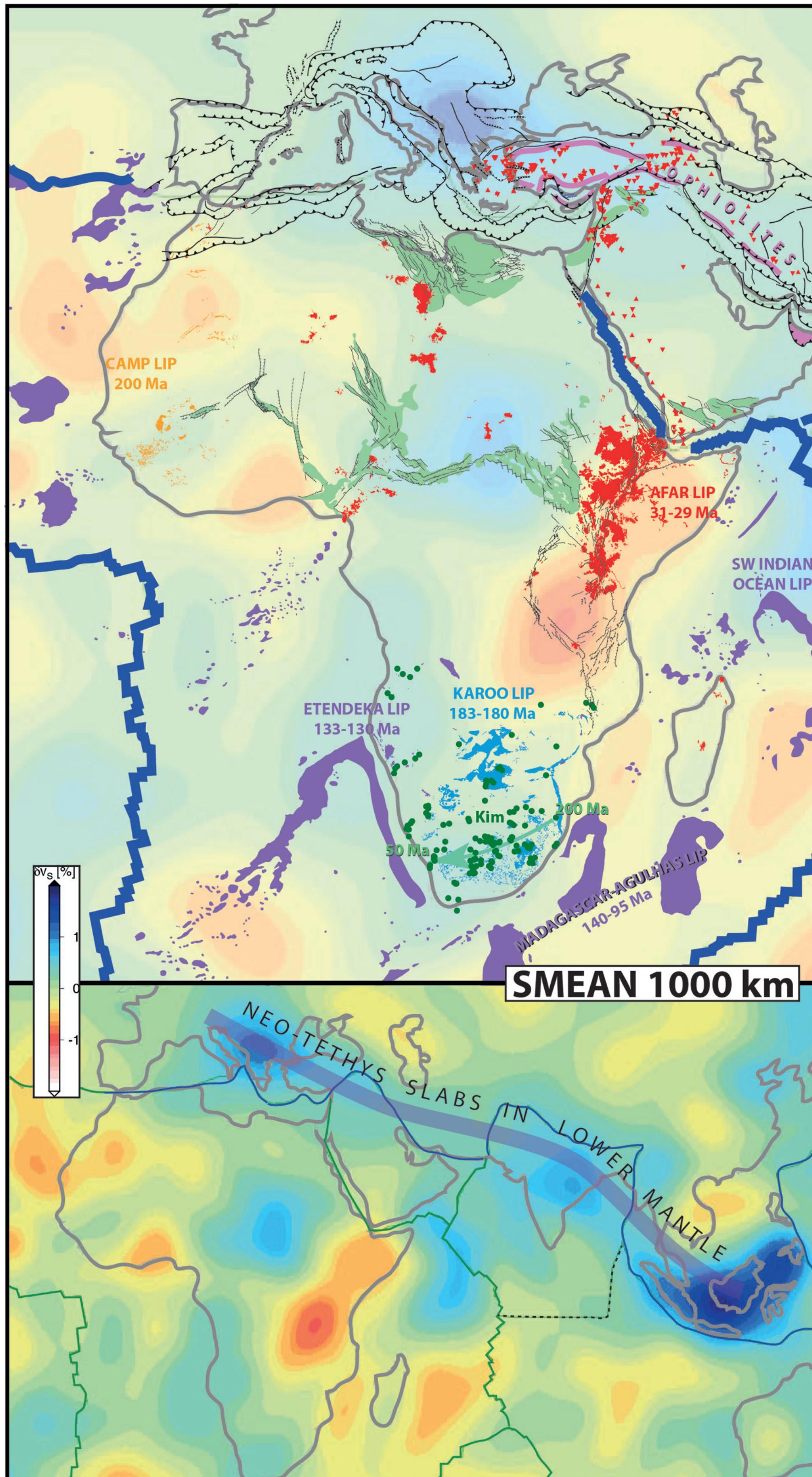
Africa from mantle plumes to subduction

During this evolution, several mantle plume events and associated large igneous provinces (LIPS) are recognized (Fig. 3), from the Central Atlantic Magmatic Province (CAMP) at 200 Ma (Marzoli et al. 1999), the Karoo event some 183–182 Ma ago (Riley et al. 2004; Svensen et al. 2012), the Etendeka LIPS between 135 and 130 Ma (Turner et al. 1994; Dodd et al. 2015), the Madagascar–Agulhas LIPS around 100 Ma within a greater southeast African LIP (Gohl et al. 2011), and finally the Afar volcanism ~45 to 30 Ma ago (Hofmann et al. 1997; Ershov and Nikishin 2004) further north and the subsequent northward migration of intraplate volcanism across the western Arabian plate and eastern Anatolia (Courtillot et al. 1999; Faccenna et al. 2013*b*; Gaina et al. 2013). The Madagascar–Agulhas LIP, in particular, has been present offshore South Africa from ~140 to 95 Ma (Gohl et al. 2011). These successive volcanic events sign the presence of a long-lasting mantle upwelling underneath South Africa during a long period, also responsible for the evolution of dynamic topography in eastern Africa (Burke 1996; Burke et al. 2008; Moucha and Forte 2011; Torsvik et al. 2014). This conclusion is further corroborated by the occurrence of kimberlites from ~200 to ~50 Ma with a younging from east to west (Jelsma et al. 2009, 2004; Torsvik et al. 2010) suggesting that South Africa has slowly overridden the plume, before the latter migrated northward (Braun et al. 2014). The plume influence is also suggested by erosion and uplift of the Southern African Plateau in the Late Cretaceous (Fig. 2, label 16; MacGregor 2010; Guillocheau et al. 2012; Colli et al. 2014). Similar uplift and erosion is also recorded in West Africa (Leprêtre et al. 2014). Frizon de Lamotte



Age (Ma)
African slab
slab penetration
India-Asia collision zone
HP Ama Drime
UHP Tso Morari
met. soles
HP-LT Indus Tsangpo
Black Sea
spreading
rifting
Iran Sistan
HP-LT
Iran Sabzevar
HP-LT
backarc spreading
HP granulites
backarc spreading
Upper plate
Zagros Blueschists
Main subduction zone
HP-LT
Armenia
HP-LT
met. soles
Turkey
HP-LT
obduction
Oman
HP-LT
oceanic accretion
Aegean
backarc opening
subduction
W. Med.
backarc opening
subduction
subduction initiation
Paris Basin folding
Isthme durancien bauxites
Africa (S. Africa margins accumulation rate)
volume 12.10^14 m^3
rifting
"Santonian event"
subsidence of North Africa
Indian plate motion
rates (cm/yr)
African plate motion (relative to Eurasia)
rates (cm/yr)
Africa slows down
Large Igneous Provinces
Afar
SW I SEA
Mad. Agh.
Pacific super swell
Parana Etendeka
Global oceanic flux (ridges-plateaus). Sea level
flux (km3/yr)
Sea level
200 m
100 m
Spreading rates
half rates (cm/yr)
South Atlantic full rate

Fig. 3. Tectonic and volcanic features of Africa and the Mediterranean and lower mantle seismic velocity anomalies (1000 km, SMEAN model (Becker and Boschi 2002)). Upper panel: Tectonic map of Africa after Milesi et al. (2010) and the Mediterranean region showing the main Cretaceous Rifts (in light green), and the main volcanic provinces (Afar in red, Southwest Indian Ocean, Madagascar–Agulhas, Etendeka, and CAMP in violet, Karoo in blue). Dark green dots are the main kimberlites (Jelsma et al. 2004, 2009). Lower panel: SMEAN tomographic model and the outline of the Neo-Tethys slab in the lower mantle after Faccenna et al. (2013*a*). [Colour online.]



et al. (2015) have recently discussed the different styles of rifting that led to the fragmentation of Gondwana, emphasizing the difference between "passive" rifting episodes (the rift is localized by inherited structures) and "active" ones (the rift is localized by the plume that weakens the lithosphere) temporally related to evidence of plume activity. This new understanding derives from modelling by Burov and Gerya (2014) showing that a plume cannot trigger a rifting without external extensional forces as it was previously supposed by Şengör and Burke (1978). The Karoo LIPS (Fig. 3) at 183 Ma is associated with an episode of rifting leading to the opening of the West Indian Ocean. Similarly, the Early Cretaceous rifting episode can be seen as a consequence of the Paraná–Etendeka LIPS.

From the Late Cenomanian until the Eocene, the northern part of Africa was under sea water and subsiding with a decrease at the Paleocene–Eocene boundary (Guiraud et al. 2005; Swezey 2009). Africa thus shows in the Late Cretaceous and early Cenozoic subsidence of the northern regions in the vicinity of the subduction and uplift in the south above mantle plumes. Convergence is accommodated by several subduction zones, one along the southern margin of Eurasia and two others where continental lithosphere underthrusts oceanic lithosphere, which all show the same asymmetry with the southern plate underthrusting the northern one.

A horizontal tomographic section (Becker and Boschi 2002) (Fig. 3) in the lower mantle (1000 km) distinctly shows the plume now sitting below eastern Africa at this depth. It also shows cold material further north, interpreted as remnants of the Neo-Tethys slab in the lower mantle (Ricard et al. 1993; Lithgow-Bertelloni and Silver 1998; Ritsema et al. 1999; Van der Voo et al. 1999; Steinberger 2000; Burke and Torsvik 2004; McNamara and Zhong 2005; Hafkenscheid et al. 2006; Garnero and McNamara 2008; van der Meer et al. 2010; Faccenna et al. 2013*a*) from the eastern Mediterranean region to India and Indonesia, signing the place where the slab has penetrated the upper–lower mantle transition zone. The timing of full penetration across the transition zone can be loosely bracketed by matching reconstructions and tomographic images (backward reconstructions of the slabs seen on tomographic models) between 70 and 45 Ma with probable significant error bars (Faccenna et al. 2013*a*; Replumaz et al. 2013).

The Eocene compression has more limited effects compared to the Late Cretaceous one, and these are mostly restricted to the western end of the convergence system with, however, the reactivation of earlier extensional structures in the east (Arsenikos et al. 2013). Nevertheless, the effects of compression are felt over a large domain from the Atlas Mountains to the Pyrenees (Frizon de Lamotte et al. 2000; Vergés et al. 2002; Mouthereau et al. 2014). This compression ended at ~35 Ma, when the subduction regime changed and back-arc extension started (Jolivet and Faccenna 2000).

Plate velocities

The two successive periods of compression (100–75 Ma and 45–35 Ma) correspond to faster convergence between Africa and Eurasia (Fig. 2, label 10). They are separated during the Paleocene by a period of very slow convergence (Fig. 2, label 10; Rosenbaum et al. 2002). The progressive build-up of Late Cretaceous compression is also coeval with an increase of the Africa absolute velocity (Fig. 2, label 11; Gaina et al. 2013), as well as a maximum of sea level at global scale (Fig. 2, label 7; Müller et al. 2008*a*, 2008*b*). It is also contemporaneous with higher velocities of spreading in the South Atlantic (Fig. 2, label 2b). The absolute motion of Africa gradually increased from the time of emplacement of the Paraná–Etendeka LIP in the southern Atlantic to the emplacement of the Madagascar–Aghulas LIP in the Indian Ocean (Fig. 2, label 11; Gaina et al. 2013). After this period, the absolute motion of Africa slowed down before a new peak before 30 Ma and a new decrease afterward. The peak of convergence velocity predates the peak of absolute velocity (Fig. 2, labels 10, 11, 13). Faster African plate motion is associated with a period of increased spreading rate between 120 and 70 Ma in the southern Atlantic (Fig. 2, labels 2, 2b; Cogné and Humler 2006; Conrad and Lithgow-Bertelloni 2007; Colli et al. 2014), although the global peak of oceanic crust production is recorded earlier at 120 Ma coeval with the Pacific superplume (Fig. 2, label 3; Larson 1991; Conrad and Lithgow-Bertelloni 2007). Subduction initiation in the Tethys Ocean in the Late Cretaceous, as shown by the age of metamorphic soles beneath ophiolites (Fig. 2, labels 21, 22, 23), temporally coincides with the period of increasing absolute velocity and the peak of convergence velocity (see green-shaded periods in Fig. 2). Similarly, the peak of Indian Plate velocity (Fig. 2, labels 14, 15) coincides with the age of metamorphic sole below ophiolites obducted onto the northern and western margins of India and the age of the southwest Indian plume at the very end of the Cretaceous (Gnos et al. 1997).

Discussion

In a recent paper, Şengör and Stock (2014) have analyzed the Late Cretaceous compressional episode along the northern margin of Africa and proposed the name of Ayyubid Orogen. In their interpretation, the eastern part of this orogen, equivalent to the Croissan Ophiolitique Péri-Arabe of Ricou (1971), results from the obduction of the Tethys oceanic floor, while the western part results from an aborted obduction. They have analyzed the kinematic changes in the motion of Africa and conclude that the Ayyubid Orogen started to form before the kink in the motion path of Africa at 84 Ma (see also Figs. 1C and 1D), which implies that some other cause should be investigated. This also shows that the state of stress in the subducting plate is not simply related to the direction of convergence between Africa and Eurasia. They further propose that one has to invoke the plates that were lost during the convergence process, but they do not put forward an explanation for the change in stress regime that finally led to the observed obduction. The data compiled here illustrate that obduction is a large-scale tectonic process completed within a short time frame. One may then speculate that it is related to a large geodynamic cause, involving changes in subduction dynamics, increase plume activity, and plate velocity increase and not only to local plate motion re-orientation. Vaughan and Scarrow (2003) proposed for instance that obduction can be linked to superplume events producing compression over the entire subduction system. The period of compression in the Late Cretaceous, including obduction, is indeed coeval with faster convergence, increasing absolute motion of Africa and interaction with superplumes. Recent studies have shown the importance of mantle plumes in governing the mantle flow beneath Africa and supporting its topography (Forte et al. 2010; Glišović et al. 2012; Moucha and Forte 2011; Glišović and Forte 2014). Large-scale mantle plumes emanating from large low-shear velocity provinces in the lower mantle, one below South Africa and one below the Pacific (Behn et al. 2004; Torsvik et al. 2006; Burke 2011; Bower et al. 2013), are thought to be stable over tens or hundreds of millions of years (Glišović et al. 2012; Bower et al. 2013), suggesting a rather stable pattern of convection in the mantle. Strain pattern in the mantle below Africa, deduced from SKS seismic anisotropy, is furthermore compatible with northward mantle flow related to the African superplume (Bagley and Nyblade 2013; Hansen et al. 2012). Similarly, the northward motion of Arabia, after its separation from Africa some 30 Ma ago, and the migration of hotspot-related volcanism toward the collision zone are also compatible with a northward asthenospheric flow dragging Africa and Arabia (Faccenna et al. 2013*b*). Plume drag and push efficiency is attested by the fact that the motion of Arabia did not stop after collision, although there is no longer any significant slab to power its northward motion (McQuarrie et al. 2003; Alvarez 2010; Faccenna et al. 2013*b*).

Recent analogues experiments (Agard et al. 2014; Edwards et al. 2015) suggest that subducting a continental margin below a denser oceanic lithosphere is feasible once subduction has initiated, which remains the main problem. In a set of models comparable to the Tethyan system, Agard et al. (2014) show that the jamming of the northern subduction can lead to subduction initiation further south, leading to obduction on the African margin, reemphasizing the model proposed by Agard et al. (2007) in which faster plate velocity renders subduction more difficult, putting the system in compression and inducing the formation of a new subduction zone.

Elaborating on the ideas suggested by Vaughan and Scarrow (2003), we now discuss a new plausible scenario coupling the evolution of the Tethyan subduction with a superplume below South Africa (Fig. 4) that should take into account three main large-scale observations: (*i*) the repeated detachment of continental pieces from Africa in the north; (*ii*) the contemporaneity of faster Africa motion, plume activity in the south, and compression and obduction; and (*iii*) the systematic southward polarity of obduction (northward subduction).

Step 1 (Fig. 4A): Assuming a continuous northward asthenospheric flow, which the superplume is part of, since the Jurassic, we first propose that extension and fragmentation of Africa (in the middle Jurassic and "middle" Cretaceous) result from this flow and the shear, or the push, it imposes to the base of the lithosphere (Bott 1993; Ziegler 1992; Stoddard and Abbott 1996). Africa was thus driven both by this drag and push and by slab pull in the subduction zone below Eurasia through a convective conveyor belt. We speculate that as long as the continuity of the conveyor belt existed, plume push resulted in extension episodes and sometimes in the stripping of microcontinents in the north of Africa. This hypothesis offers the advantage of explaining the repetition of the same process above a continuous northward mantle flow from the Paleozoic to the present, if Africa has been all along under the influence of mantle flow largely controlled by a long-lasting plume. Before plate acceleration and compression, the slab was not restricted to the upper mantle and partly retreated southward, leading to back-arc rifting in the upper Eurasian plate (Black Sea; South Caspian; Nain-Baft, Sabzevar, and possibly Sistan in central Iran).

Step 2 (Fig. 4B): Obduction. Interactions of mantle flow and cratonic lithospheric keels has long been discussed (Stoddard and Abbott 1996), and it has been recently suggested that continental cratonic keels may lead to plate acceleration when a plume arrives (Zahirovic et al. 2015). The detailed interactions between a plume and a continent above are not known but some recent investigations by Koptev et al. (2015) have confirmed the early work of Stoddard and Abbott (1996) and the importance of the push of the plume of the irregularities of the base of the lithosphere to move continents. Following up on this suggestion, we propose that the push due to the African superplume forced the (African) slab beneath Eurasia into the upper mantle at a faster rate. Slab penetration in the lower mantle could have produced a surge of compression within the subducting plate, leading to the formation of a new subduction zone close to the North African margin (Figs. 4B). Stresses would then build up until the whole system is in compression, leading to the so-called Santonian event. The Late Cretaceous compression is associated with an uplift of South Africa when it passes above the plume, while North Africa is instead subsiding and subducting below the obducting Tethys oceanic lithosphere. A mechanical link between slab penetration and the progressive reorientation of the Africa–Eurasia convergence around 84 Ma should be studied.

Step 3 (Fig. 4C): These compressional stresses would then be relaxed once the penetration beneath Eurasia is complete in the lower mantle and the conveyor belt is restored, that is some 70 Ma ago, when the Neo-Tethys obduction process stopped. The 25 Myr of compression would then be the time needed from plate acceleration to full slab penetration into the lower mantle. The same process may have happened again further east at the very end of the Cretaceous with the obduction of ophiolites on top of the northern and western margins of India. It may also have happened earlier at the end of the Jurassic to form the Dinaric ophiolite.

The boundary conditions of the model would then be radically different before and during the compressional episode. Before compression the plume in the south pushes on a plate that is free to move northward on its northern boundary and the system is driven by slab pull that is more efficient than plume push. During compression the more active plume induces a stronger push in the south while the slab is penetrating the upper-lower mantle discontinuity in the north, thus inducing compression in the whole northern half of the African plate.

This scenario can furthermore create the asymmetrical boundary conditions explaining the polarity of the newly created subductions and the underthrusting of old and dense continental mantle below younger and lighter oceanic mantle. We show in Fig. 5 a possible evolution at the start of obduction, shortly after a young oceanic domain had formed (hence before compression started). The mantle flow dragging, or pushing, Africa northward imposes a general shear sense such that the newly created subduction is dipping north, leading to the underthrusting of the African margin and continental domain, supported by an old and cold lithospheric mantle, below the young and hot oceanic domain newly formed.

Step 4 (Fig. 4D): In the eastern part of the system the conveyor belt is active between the slab penetrating deep in the mantle in the north and the plume pushing the African plate toward the north. The situation is quite similar to the Jurassic stage with this time extension active in the future Gulf of Aden and Red Sea and the separation of Arabia from Africa. Back-arc extension is active in the Mediterranean. In the west, compression resumed in the Eocene in Africa and the most significant compressional tectonics is restricted to the western half of the system, especially in the Atlas system, with also some inversion of extensional features farther east from Cyrenaica to the Palmyrides. This new compressional period is coeval with the building of the Hellenides–Taurides accretionary wedge at the expense of the Pindos Ocean and then the northern margin of Apulia. In the westernmost Mediterranean, compression is active from the Atlas to the Pyrenees and the question of stress propagation from Africa to Eurasia must be discussed. If collision was already going on in the future Gibraltar Arc (Jolivet and Faccenna 2000; Jolivet et al. 2003), compressional stresses were transmitted across the collision zone within the lithosphere from the Atlas to the Pyrenees. If one assumes instead that collision was not yet effective and that some narrow oceanic domain was still present between Africa and Iberia (Vergés and Sabat 1999; Frizon de Lamotte et al. 2000), the compressional stress regime before 35 Ma may be due to the progressive formation of the subduction zone since the Late Cretaceous, very slow in the first period and then becoming mature enough for arc volcanism to develop and slab pull to be active and accommodate slab retreat after 35–30 Ma. The slight velocity increase of the absolute motion of Africa seen in some kinematic models between 40 and 30 Ma may also suggest that coupling with the Afar plume when it impacted the base of the African lithosphere has accelerated the displacement of Africa before the effective collision in the Caucasus–Zagros region that slowed it drastically.

An alternative scenario was recently discussed by Jagoutz et al. (2015) to explain the velocity increase of the India–Asia convergence in the Latest Cretaceous, involving a double subduction north of India. The southernmost of these two subduction zones in the west is equivalent to the obduction zone in our model. So the geometry is similar in the two models, and Jagoutz et al. (2015) argue through numerical modelling that this double subduction

Fig. 4. Scenario linking the Late Cretaceous obduction of the Tethyan ophiolites on Apulia and Africa to convection. (A) ~170 Ma. An established conveyor belt with plume push and slab pull below Africa drags continental blocks northward away from the main African plate, forming the Apulia microcontinent. The whole system is under extension, rifts develop far within the African plate (Late Jurassic – Early Cretaceous). (B) ~90 Ma. The slab has detached below Eurasia and the African plate is driven by plume push and less slab pull. The plume beneath South Africa is more active and accelerates the northward drift of Africa and pushes the African slab faster in the mantle thus rendering subduction more difficult and putting the whole system under compression. Compression during penetration of the slab in the lower mantle induces shortening within Africa and Eurasia and initiates a new subduction along the northern margin of Africa, leading to obduction of the oceanic crust onto the continental margin (100–90 Ma). At maximum compression shortening propagates far inside Africa reactivating all the Cretaceous rifts (Santonian event, 85 Ma). (C) ~70 Ma. Slab avalanches in the lower mantle and a progressive relaxation of compressional stresses ensues. (D) ~30 Ma. After full penetration, the conveyor belt is functional again. The slab is deep into the lower mantle like below the Aegean and retreats. Back-arc extension in the upper plate is observed. Extension is active above the northward moving plume leading to the separation of Arabia from the main body of the African plate. [Colour online.]

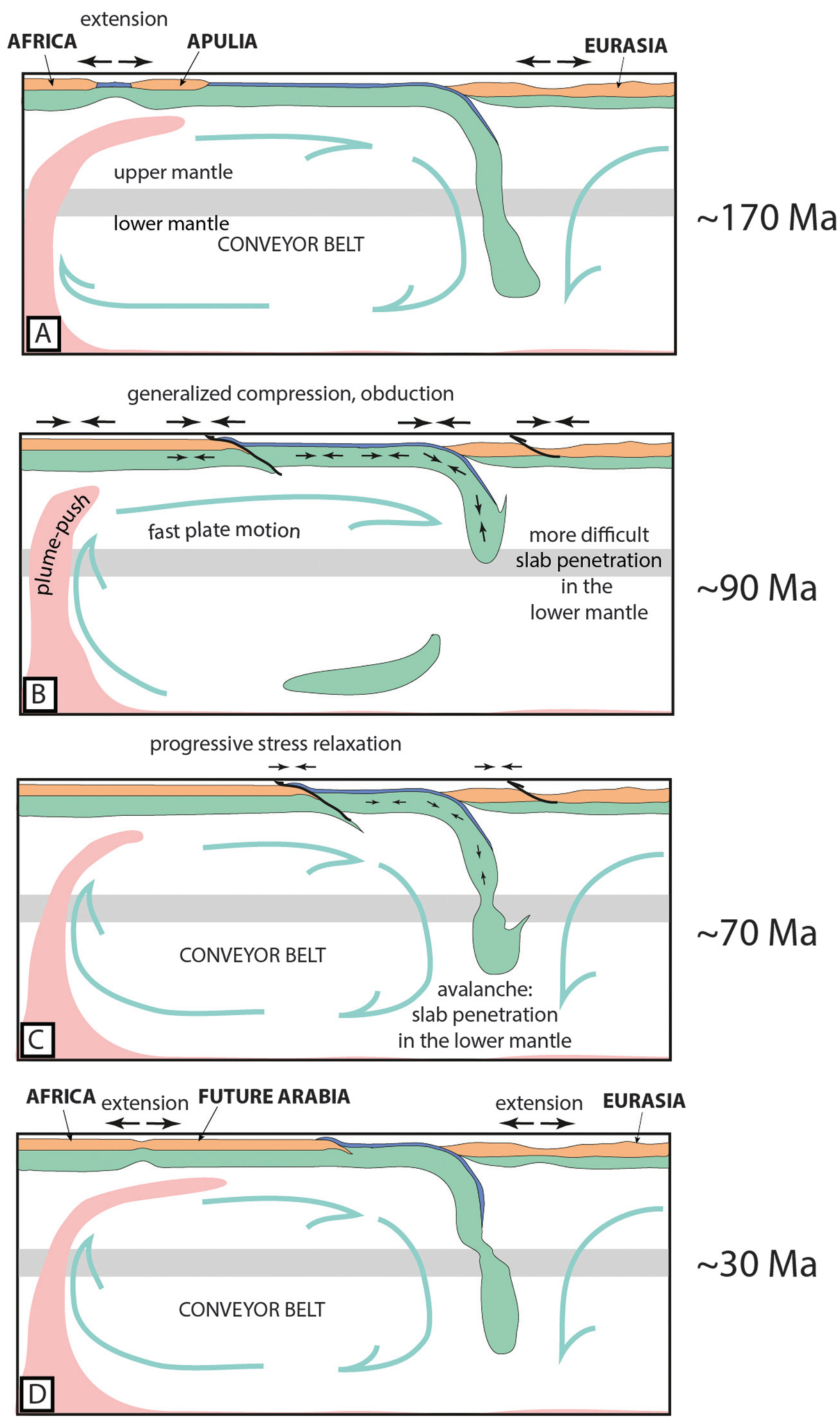
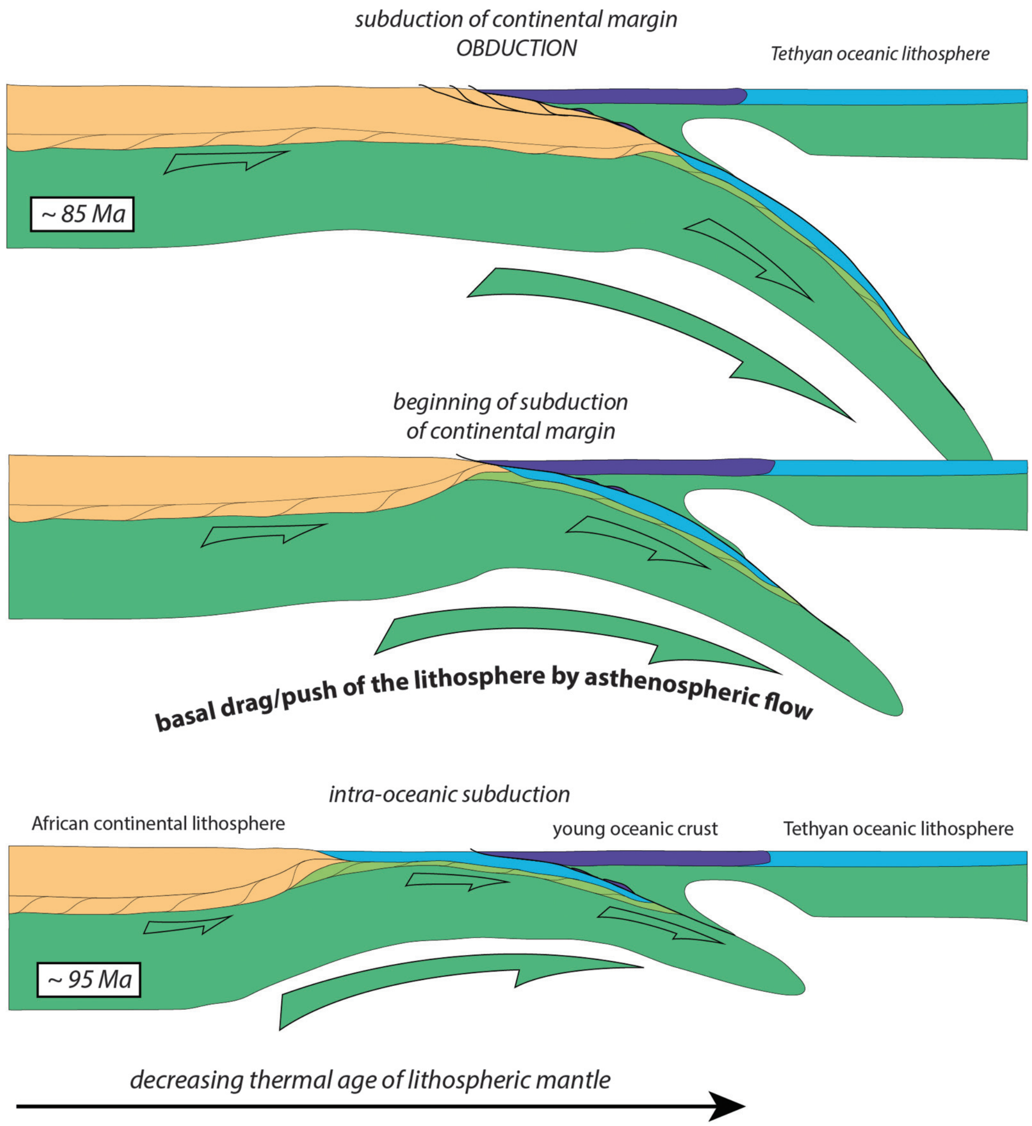


Fig. 5. A scenario of obduction driven by basal drag. Northward movement of the asthenospheric mantle with respect to Africa favours the subduction of the subcontinental lithospheric mantle below the young and light oceanic one. [Colour online.]



is likely to have accelerated convergence, but they do not address the question of the initiation of this subduction zone, which is the topic of the present paper, nor do they explain the observed polarity of subduction.

Conclusion

Our scenario aims at integrating large-scale tectonic processes within a progressive evolution of the convergence zone between Africa and Eurasia. Because the scale of plate fragmentation and subsequent obduction is so large, we look for processes involving the whole mantle. We stress that the Late Cretaceous compression cannot have been a consequence of continental collision since the Neo-Tethys Ocean was still widely open when it happened. We propose that the alternation of periods of extension and compression (including obduction) in Africa and the Neo-Tethys realm result from changes in the dynamics of convection underneath, combined with regional-scale reorganizations. The long-term geological record suggests that the normal extensional situation corresponds to steady-state whole-mantle convection, Africa being carried northward by a large-scale conveyor belt, through plume push in the south and slab pull in the north, while the Late Cretaceous compression and obduction would result from plate acceleration due to increasing plume push below South Africa and difficult penetration before slab avalanching in the lower mantle, until full penetration of the Tethyan slab across the transition zone and reestablishment of the conveyor belt. This would have produced a strong compression at plate boundaries and, as a consequence, would have activated a new plate boundary at the location of previous crustal weakness (Agard et al. 2014). The asymmetry of the northward mantle flow may explain the polarity of initiation of a northward subduction, leading to obduction with oceanic crust emplaced on top of the subducting continental margins of Africa and Apulia through basal drag or push. Although the scale of the obduction at the Jurassic–Cretaceous boundary in the Dinarides cannot compare with the Late Cretaceous one, a similar scenario could be envisaged. The later Eocene compression would be the consequence of either subduction initiation in the western Mediterranean or plate collision. As superplumes and subduction zones, which govern mantle convection, are long-lived features, testing models of mantle convection requires using sequences of events on long durations that only the geological record is able to provide. Such scenarios can be useful to modellers as they may provide time constants (~20 Myr) for the establishment and destruction of whole-mantle convection cells or slab penetration in the lower mantle.

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